Comparison of seasonal terrestrial water storage variations from GRACE with groundwater-level measurements from the High Plains Aquifer (USA)

Gil Strassberg,[1] Bridget R. Scanlon,[1] and Matthew Rodell[2]

Received 25 March 2007; revised 18 May 2007; accepted 6 June 2007; published 18 July 2007.

[1] This study presents the first comparison of seasonal groundwater storage (GWS) variations derived from GRACE satellite data with groundwater-level measurements in the High Plains Aquifer, USA (450,000 km^2). Correlation between seasonal GRACE terrestrial water storage (TWS) and the sum of GWS estimated from field measurements (2,700 wells) and soil moisture (SM) simulated by a land surface model is high (R = 0.82). Correlation between GRACE-derived and measured GWS is also significant (R = 0.58). Seasonal GRACE-derived TWS and GWS changes were detectable ($\geq$ uncertainty) in 7 and 5 out of 9 monitored periods respectively whereas maximum changes (between winter/spring and summer/fall) in TWS and GWS were detectable in all 5 monitored periods. These results show the potential for GRACE to monitor GWS changes in semiarid regions where irrigation pumpage causes large seasonal GWS variations. **Citation:** Strassberg, G., B. R. Scanlon, and M. Rodell (2007), Comparison of seasonal terrestrial water storage variations from GRACE with groundwater-level measurements from the High Plains Aquifer (USA), *Geophys. Res. Lett.*, *34*, L14402, doi:10.1029/2007GL030139.

1. Introduction

[2] Water scarcity is a critical issue, as an estimated 1.1 billion people worldwide lack access to clean drinking water [*World Health Organization*, 2003]. Water scarcity is greatest in semiarid regions, which occupy ~30% of the Earth's land [*Dregne*, 1991; *Vorosmarty et al.*, 2000]. Irrigated agriculture is the primary consumer of global freshwater resources, accounting for an average of ~90% of freshwater consumption during the past century [*Shiklomanov*, 2000]. Groundwater (GW) based irrigation systems have expanded greatly during the past several decades, particularly in the North China Plain and western India, resulting in large-scale aquifer depletion [*Scanlon et al.*, 2007]. Monitoring GW levels in these regions is extremely limited.

[3] The Gravity Recovery and Climate Experiment (GRACE), launched in March 2002, provides monthly measurements of Earth's gravity field [*Tapley et al.*, 2004]. Over land, and accounting for atmospheric circulation, changes in the gravity field are mainly attributed to variations in terrestrial water storage (TWS), which is a vertically integrated measure of water storage that includes GW, soil moisture (SM), surface water, snow water, and vegetation water. Hence GRACE gravity data can be used to infer temporal variations in TWS. Previous studies have shown that water-storage variations derived from GRACE compare favorably with those based on land surface models and combined atmospheric-terrestrial water balances [*Rodell et al.*, 2004; *Syed et al.*, 2005; *Hu et al.*, 2006; *Niu and Yang*, 2006] and that TWS can be used to estimate changes in components of the water budget (e.g. total basin discharge, GWS, SM, and ET). Comparisons of GRACE TWS and GWS estimated as TWS-SM with monitored SM and GWS for Illinois and Oklahoma [*Yeh et al.*, 2006; *Swenson et al.*, 2006; S. Swenson et al., Estimating profile soil moisture and groundwater storage variations in the southern Great Plains using GRACE satellite gravimetric and Oklahoma Mesonet soil moisture data, submitted to *Water Resources Research*, 2007] and with simulated SM and monitored GWS in the Mississippi basin [*Rodell et al.*, 2006] also showed good agreement.

[4] The High Plains aquifer has been proposed as an ideal location to test GRACE TWS changes since the inception of GRACE because of the aquifer area (450,000 km^2), and because it is the most intensively monitored aquifer at this scale globally [*U.S. National Research Council Committee on Earth Gravity from Space*, 1997]. The aquifer is unconfined with saturated thickness from 0 to 300 m (mean 60 m) and water-table depth from 0 to 150 m (mean 30 m) [*Dennehy*, 2000]. Extensive GW-monitoring campaigns have been conducted annually since 1988 to produce GW-level maps and to estimate GWS changes (~9,200 wells monitored in 2003 [*McGuire*, 2004]).

[5] The climate of the region is semiarid, with annual precipitation (P) from 400 to 600 mm and mean annual pan evaporation (PE) from 1,500 to 2,700 mm. The low P/PE ratio results in large agricultural areas (50,000 km^2, ~30% of cropland) being dependent on GW irrigation (Figure 1a). Extensive irrigation (30% of GW used for irrigation in the US [*Dennehy*, 2000]) decreased GW levels by an average of ~4 m over the entire aquifer since predevelopment (~1950) and by up to 68 m in some regions [*McGuire*, 2004]. About 90% of GW withdrawals occur during the summer [*Amosson et al.*, 2003].

[6] This study evaluates the use of GRACE TWS and modeled SM to estimate seasonal changes in GWS in the High Plains aquifer. This represents the first comparison of GRACE data with GW-level measurements in a semiarid

[1]Bureau of Economic Geology, Jackson School of Geosciences, The University of Texas at Austin, Austin, Texas, USA.
[2]Hydrological Sciences Branch, NASA Goddard Space Flight Center, Greenbelt, Maryland, USA.

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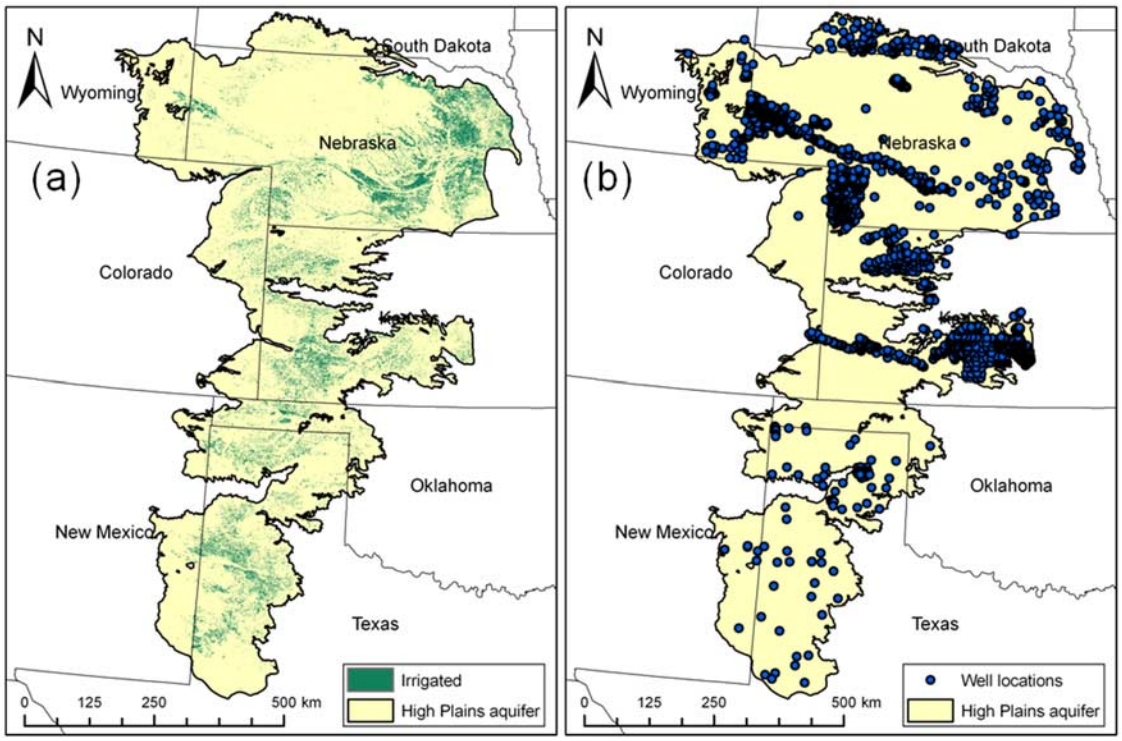

Figure 1. (a) Location of irrigated areas over the High Plains aquifer [*Qi et al.*, 2002] and (b) location of wells where seasonal water-level changes were calculated (total of 2,719 wells).

region where large-scale irrigation pumpage affects seasonal GWS variations.

2. Data and Methods

[7] Variations in TWS include a vertically integrated measure of water-storage changes in GW, SM, surface water, snow, and biomass. Most surface water in the High Plains is internally drained into ephemeral lakes or playas (~53,000) representing ~0.5% of the land surface area, with few large river systems. Based on analysis of playa wetted area from Landsat images (300 images between 1985 and 2000 [*Howard et al.*, 2003]), estimated maximum variation in playa water storage was ≤3.5 mm. Simulated monthly mean snow water equivalent (see section 2.3) ranged from 0 to 2.1 mm (mean 0.25 mm) over the study time period (2003–2005), and estimated seasonal biomass changes were <5 mm [*Rodell et al.*, 2005]. *Rodell and Famiglietti* [2002] estimated that interannual GWS and SM changes in the High Plains averaged 20 and 24 mm, respectively. Seasonal GWS and SM changes are expected to exceed interannual variations, suggesting that GWS and SM are the dominant components in TWS variability:

$$\Delta TWS = \Delta GWS + \Delta SM \tag{1}$$

where Δ refers to change. Seasonal GWS changes were estimated by reorganizing equation (1) (ΔGWS = ΔTWS − ΔSM) and results were compared to GWS changes derived from GW-level measurements.

2.1. GRACE TWS

[8] The University of Texas Center for Space Research (CSR), GeoFoschungsZentrum Potsdam (GFZ), and Jet Propulsion Laboratory (JPL) each produce time series of surface mass in spherical harmonic format based on GRACE intersatellite range rate data. These data were processed with a 400-km radius Gaussian smoother and destriped to remove spurious north-south trending bands which appear in the raw fields [*Swenson and Wahr*, 2006] and averaged following the recommendation of *Chambers* [2006]. A total of 30 approximately-monthly TWS measurements between January 2003 and December 2005 were used in the analysis (excluding missing data from June 2003, January 2004, and July–October 2004). Seasonal TWS anomalies were calculated by averaging monthly anomalies over 3-month periods (January–March, April–June, July–September, and October–December).

2.2. Groundwater Storage

[9] Seasonal GWS variations were calculated from GW levels measured throughout the aquifer. GW-level data were obtained from the USGS (http://waterdata.usgs.gov), Texas Water Development Board (http://www.twdb.state.tx.us), and Kansas Geological Survey (www.kgs.ku.edu). Mean seasonal GW levels were calculated for each well and GW-level changes were calculated as the difference between GW levels in two consecutive seasons. Seasonal GW-level changes from January 2003 to December 2005 were calcu-

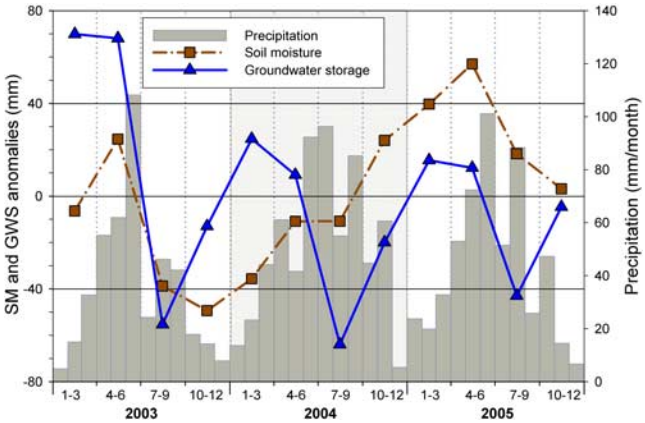

Figure 2. Seasonal measured GWS, simulated SM, and monthly precipitation for the High Plains aquifer. GWS and SM data are shown as anomalies relative to the mean for the analysis period (2003–2005) and units represent equivalent thickness of water (mm).

lated based on GW levels from 2,719 wells (Figure 1b). The number of available GW-level data for each season varied from 657 to 1,891 (mean 1,050 wells per season). Many of the wells are proximal to irrigated areas (1194 wells within 100 m of irrigated fields); therefore, to eliminate errors associated with water levels in active pumping zones or measurement errors, data were filtered to exclude extreme changes. Daily GW-level records from 75 wells in the USGS database showed that the seasonal amplitude of GW levels is $\leq$4.6 m in 97% of wells. Thus, GW-level changes >4.6 m were excluded from the analysis. GW-level changes were regionalized to the area of the High Plains by calculating a spatial mean. Point data were averaged over a 1 $\times$ 1 degree mesh where each cell of the mesh was assigned the mean value of all wells within it. An average GW-level change was calculated using an area-weighted scheme, where the value in each cell was multiplied by the area of the cell overlapping the High Plains aquifer. Specific yield is spatially heterogeneous ranging from less than 0.1 to 0.3, with an average of 0.15 [*Gutentag et al.*, 1984]. GWS changes were calculated by multiplying seasonal GW-level changes by the average specific yield (0.15).

2.3. Soil Moisture

[10] Variations in SM were estimated with results from the Noah land surface model [*Ek et al.*, 2003]. Forcing and parameterization of the model are based on data from the North American Land Data Assimilation System (NLDAS), which includes hourly observation-based precipitation and solar radiation, among other meteorological fields, and high quality soil and vegetation parameter fields, on a 1/8 degree grid over central North America [*Cosgrove et al.*, 2003]. Simulated results included SM in the top 2 m of the soil column, snow water equivalent, and canopy interception. The results were averaged over 3-month periods to derive seasonal changes. Changes in SM accounted for over 99% of the seasonal mass change; therefore, snow water equivalent and canopy interception were omitted from further analysis.

2.4. Uncertainty

[11] Uncertainties in monthly GRACE TWS anomalies were estimated by fitting a function (annual and semiannual cycles + trend) to the averaged TWS anomalies and calculating the root mean square deviation (RMSD) between measured TWS anomalies and the fitted function. The RMSD between TWS data and the fitted function represents a conservative estimate of uncertainty in TWS measurements because it assumes that deviations result only from measurement error and not from actual TWS variations. Uncertainties in averaged seasonal anomalies were estimated from monthly uncertainties

$$\delta_N = \delta_1/\sqrt{N} \tag{2}$$

where δ_N is uncertainty in an anomaly averaged over N months, and δ_1 is uncertainty in a 1-month anomaly. The RMSD between monthly TWS anomalies and the fitted function is 13 mm and uncertainty in seasonal TWS anomalies is 8 mm. Seasonal changes in TWS were calculated by differencing two seasonal anomalies (e.g. winter TWS anomaly – summer TWS anomaly) and uncertainty in the change was estimated as:

$$\delta_T = \sqrt{\delta_i^2 + \delta_j^2} \tag{3}$$

where δ_T is combined uncertainty and δ_i and δ_j are seasonal uncertainties. Using a seasonal uncertainty of 8 mm in equation (3) results in an uncertainty of 11 mm for TWS changes.

[12] Uncertainty in SM changes was calculated by applying a coefficient of variation (CV) to a mean SM change. Based on results from 10 land surface schemes over the High Plains, *Rodell and Famiglietti* [2002] suggested a

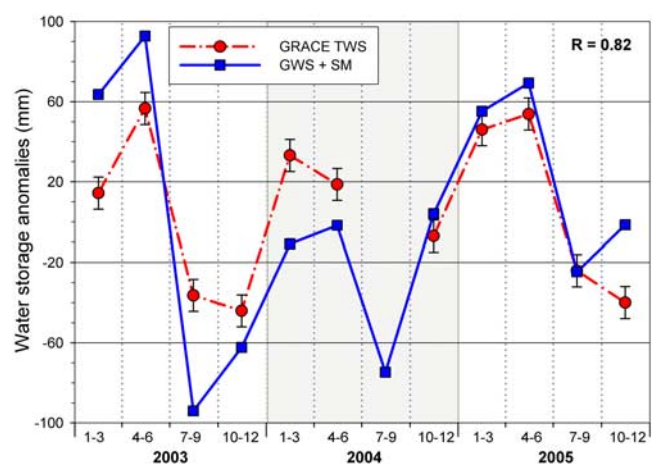

Figure 3. GRACE-derived TWS and combined GWS (from GW-level measurements) and SM (simulated) for the High Plains aquifer. Data are shown as anomalies relative to the mean for the analysis period (2003–2005) and units represent equivalent thickness of water (mm). Error bars represent TWS uncertainties.

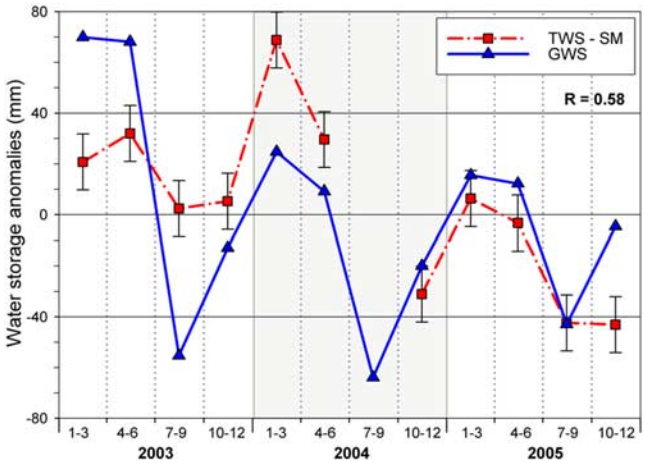

Figure 4. Seasonal GW storage variations for the High Plains aquifer derived from field measurements (GWS) and GRACE TWS and simulated SM (TWS-SM). Data are shown as anomalies relative to the mean for the analysis period (2003–2005) and units represent equivalent thickness of water (mm). Error bars represent TWS-SM uncertainties.

reasonable CV for simulated SM changes of 0.3. Applying this CV to the mean 6-month SM change (37 mm) during the analysis period yields an uncertainty of 11 mm. Using equation (3) uncertainty in SM anomalies was back calculated (assuming SM uncertainties are equal in both seasons between which a change is calculated) and is 8 mm. Applying equation (3) to TWS and SM uncertainties gives an uncertainty of 11 mm in seasonal TWS-SM anomalies and an uncertainty of 15 mm in seasonal TWS-SM changes.

3. Results and Discussion

[13] Measured GWS varied seasonally, as a result of irrigation pumpage, with highs in winter and lows in summer (Figure 2). GWS and SM anomalies were of similar magnitude with maxima of 70 mm and 57 mm, respectively. SM did not show any systematic seasonal cycle over the 2003–2005 period, which can be explained by elevated precipitation from September to November 2004 (175% of long-term mean) that caused accumulation of SM during 2004 and into 2005. The different patterns observed for SM and GWS can be explained by the disconnect between the upper part of the soil column and the GW table because of the thick unsaturated zone, and by the effect of GW pumpage for irrigation.

[14] GRACE TWS anomalies are in good agreement with combined GW and SM anomalies ($R = 0.82$, $p < 0.005$) (Figure 3). Both time series show a strong seasonal cycle, with maximum storage in winter and spring and minimum storage in summer and fall. Maximum TWS anomaly was 57 mm, and combined GWS + SM had greater variability with a maximum anomaly of 94 mm. The RMSD between the two estimates (TWS and GWS + SM) is 33 mm. Some of this difference between the two estimates can be explained by overestimation of GWS variations during summer because many monitored wells are proximal to irrigated areas.

[15] To evaluate detectability of the TWS signal, TWS changes were compared to the uncertainty. Seasonal changes, calculated as the difference between two consecutive seasonal anomalies, range from 8 to 93 mm (mean 43 mm). These changes are larger than the estimated uncertainty (11 mm) and are considered detectable in 7 out of 9 periods (changes were calculated for only 9 out of 11 periods due to missing TWS data). Maximum TWS changes (between winter/spring and summer/fall) range from 40 to 101 mm (mean 75 mm), much larger than the estimated uncertainty (11 mm) and are considered detectable in all 5 monitored periods.

[16] Seasonal GW storage calculated by subtracting simulated SM from GRACE TWS (TWS-SM) compared favorably with GWS derived from GW-level measurements (Figure 4). Both have similar magnitudes and show strong seasonal trends with storage increasing in winter and decreasing in summer ($R = 0.58$, $p = 0.06$). The RMSD between the two estimates (GWS and TWS-SM) is 33 mm. Seasonal TWS-SM changes range from 1 to 63 mm (mean 26 mm), exceeding the uncertainty (15 mm) and considered detectable in 5 out of the 9 monitored periods. Maximum TWS-SM changes are between 30 and 100 mm (mean 57 mm), exceeding the uncertainty and considered detectable in all 5 monitored periods.

4. Conclusions

[17] This study presents the first direct comparison of variations in seasonal GWS derived from GRACE TWS and simulated SM with GW-level measurements in a semiarid region. Results showed that variations in GWS and SM are the main sources controlling TWS changes over the High Plains, with negligible storage changes from surface water, snow, and biomass.

[18] Seasonal variations in GRACE TWS compare favorably with combined GWS from GW-level measurements (total 2,700 wells, average 1,050 GW-level measurements per season) and simulated SM from the Noah land surface model ($R = 0.82$, RMSD = 33 mm). Estimated uncertainty in seasonal GRACE-derived TWS is 8 mm, and estimated uncertainty in TWS changes is 11 mm. Estimated uncertainty in SM changes is 11 mm and combined uncertainty for TWS-SM changes is 15 mm. Seasonal TWS changes are detectable in 7 out of 9 monitored periods and maximum changes within a year (e.g. between winter and summer) are detectable in all 5 monitored periods.

[19] Grace-derived GWS calculated from TWS-SM generally agrees with estimates based on GW-level measurements ($R = 0.58$, RMSD = 33 mm). Seasonal TWS-SM changes are detectable in 5 out of the 9 monitored periods and maximum changes are detectable in all 5 monitored periods.

[20] Good correspondence between GRACE data and GW-level measurements from the intensively monitored High Plains aquifer validates the potential for using GRACE TWS and simulated SM to monitor GWS changes and aquifer depletion in semiarid regions subjected to intensive irrigation pumpage. This method can be used to monitor regions where large-scale aquifer depletion is ongoing, and in situ measurements are limited, such as the North China Plain or western India. This potential

should be enhanced by future advances in GRACE processing, which will improve the spatial and temporal resolution of TWS changes, and will further increase applicability of GRACE data for monitoring GWS.

References

Amosson, S., T. Marek, L. New, F. Bretz, and L. Stewart (2003), Estimated irrigation demand for the southern Ogallala GAM, in *GW Availability of the Southern Ogallala Aquifer in Texas and New Mexico: Numerical Simulations Through 2050*, contract report, appendix B, edited by N. T. Blandford et al., Tex. Water Dev. Board, Austin, Tex.

Chambers, D. P. (2006), Evaluation of new GRACE time-variable gravity data over the ocean, *Geophys. Res. Lett.*, *33*, L17603, doi:10.1029/2006GL027296.

Cosgrove, B. A., et al. (2003), Real-time and retrospective forcing in the North American Land Data Assimilation System (NLDAS) project, *J. Geophys. Res.*, *108*(D22), 8842, doi:10.1029/2002JD003118.

Dennehy, K. (2000), High Plains regional ground-water study, *U.S. Geol. Surv. Fact Sheet*, *FS-091-00*, 6 pp.

Dregne, H. E. (1991), Arid land degradation—A results of mismanagement, *Geotimes*, *36*(6), 19–21.

Ek, M. B., K. E. Mitchell, Y. Lin, E. Rogers, P. Grunmann, V. Koren, G. Gayno, and J. D. Tarpley (2003), Implementation of Noah land surface model advances in the National Centers for Environmental Prediction operational mesoscale Eta model, *J. Geophys. Res.*, *108*(D22), 8851, doi:10.1029/2002JD003296.

Gutentag, E. D., F. J. Heimes, N. C. Krothe, R. R. Luckey, and J. B. Weeks (1984), Geohydrology of the High Plains aquifer in parts of Colorado, Kansas, Nebraska, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming, *U.S. Geol. Surv. Prof. Pap.*, *1400-B*, 66 pp.

Howard, T., G. Wells, L. Prosperie, R. Petrossain, H. Li, and A. Thapa (2003), Characterization of playa basins on the High Plains of Texas, *Tex. Water Dev. Board. Rep. 357*, Austin, Tex.

Hu, X. G., J. L. Chen, Y. H. Zhou, C. Huang, and X. H. Liao (2006), Seasonal water storage change of the Yangtze River basin detected by GRACE, *Sci. China, Ser. D*, *49*(5), 483–491, doi:10.1007/s114300-006-0483-5.

McGuire, V. L. (2004), Water-level changes in the High Plains aquifer predevelopment to 2003 and 2002 to 2003, *U.S. Geol. Surv. Fact Sheet*, *2004–3097*.

Niu, G.-Y., and Z.-L. Yang (2006), Assessing a land surface model's improvements with GRACE estimates, *Geophys. Res. Lett.*, *33*, L07401, doi:10.1029/2005GL025555.

Qi, S. L., A. Konduris, D. W. Litke, and J. Dupree (2002), Classification of irrigated land using satellite imagery, the High Plains aquifer, nominal date 1992, *U.S. Geol. Surv. Water Resour. Invest. Rep.*, *02-4236*.

Rodell, M., and J. S. Famiglietti (2002), The potential for satellite-based monitoring of GW storage changes using GRACE: the High Plains aquifer, central US, *J. Hydrol.*, *263*(1–4), 245–256.

Rodell, M., J. S. Famiglietti, J. Chen, S. I. Seneviratne, P. Viterbo, S. Holl, and C. R. Wilson (2004), Basin scale estimates of evapotranspiration using GRACE and other observations, *Geophys. Res. Lett.*, *31*, L20504, doi:10.1029/2004GL020873.

Rodell, M., B. F. Chao, A. Y. Au, J. S. Kimball, and K. C. McDonald (2005), Global biomass variation and its geodynamic effects: 1982–98, *Earth Interact.*, *9*, 1–19.

Rodell, M., J. Chen, H. Kato, J. S. Famiglietti, J. Nigro, and C. R. Wilson (2006), Estimating groundwater storage changes in the Mississippi River basin (USA) using GRACE, *Hydrogeol. J.*, *15*(1), 159–166, doi:10.1007/s10040-006-0103-7.

Scanlon, B. R., I. Jolly, M. Sophocleous, and L. Zhang (2007), Global impacts of conversions from natural to agricultural ecosystems on water resources: Quantity versus quality, *Water Resour. Res.*, *43*, W03437, doi:10.1029/2006WR005486.

Shiklomanov, I. A. (2000), Appraisal and assessment of world water resources, *Water Int.*, *25*(1), 11–32.

Swenson, S., and J. Wahr (2006), Post-processing removal of correlated errors in GRACE data, *Geophys. Res. Lett.*, *33*, L08402, doi:10.1029/2005GL025285.

Swenson, S., P. J.-F. Yeh, J. Wahr, and J. Famiglietti (2006), A comparison of terrestrial water storage variations from GRACE with in situ measurements from Illinois, *Geophys. Res. Lett.*, *33*, L16401, doi:10.1029/2006GL026962.

Syed, T. H., J. S. Famiglietti, J. Chen, M. Rodell, S. I. Seneviratne, P. Viterbo, and C. R. Wilson (2005), Total basin discharge for the Amazon and Mississippi River basins from GRACE and a land-atmosphere water balance, *Geophys. Res. Lett.*, *32*, L24404, doi:10.1029/2005GL024851.

Tapley, B. D., S. Bettadpur, J. C. Ries, P. F. Thompson, and M. M. Watkins (2004), GRACE measurements of mass variability in the Earth system, *Science*, *305*(5683), 503–505.

U.S. National Research Council Committee on Earth Gravity from Space (1997), *Satellite Gravity and the Geosphere: Contributions to the Study of the Solid Earth and its Fluid Envelope*, Natl. Acad. Press, Washington, D. C.

Vorosmarty, C. J., P. Green, J. Salisbury, and R. B. Lammers (2000), Global water resources: Vulnerability from climate change acid population growth, *Science*, *289*(5477), 284–288.

World Health Organization (2003), *Health Human Rights Publ. Ser.*, vol. 3, New York.

Yeh, P. J.-F., S. C. Swenson, J. S. Famiglietti, and M. Rodell (2006), Remote sensing of groundwater storage changes in Illinois using the Gravity Recovery and Climate Experiment (GRACE), *Water Resour. Res.*, *42*, W12203, doi:10.1029/2006WR005374.

M. Rodell, Hydrological Sciences Branch, NASA Goddard Space Flight Center, Greenbelt, MD 20771, USA.

B. R. Scanlon and G. Strassberg, Bureau of Economic Geology, Jackson School of Geosciences, The University of Texas at Austin, Austin, TX 78758, USA.